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Paleolimnological evidence for the onset and termination of glacial aridity from Lake Tanganyika, Tropical East Africa

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Abstract

Geochemical and sedimentological data in a continuous 60,000-year sediment core record from the Kalya horst region of central Lake Tanganyika provide a detailed history of paleoclimate-mediated weathering and overflow events from upstream Lake Kivu. Univariate (elemental profiles), bivariate (elemental ratios) and multivariate analyses of chemical trends show variations between the dry Late Pleistocene (32–18 ka cal yr BP) and the wetter conditions that both preceded and post-date that interval. This record places important new constraints on the timing of aridity in East Africa during the high-latitude Last Glacial Maximum (LGM) based on significant decreases in magnetic susceptibility and soluble cation concentrations coinciding with biogenic silica. Elemental indicators in the oldest portion of the sedimentary record (60–50 ka cal yr BP) characterize this interval as a comparatively wet period, similar to modern conditions. Our record demonstrates that the ensuing transition toward arid conditions in tropical Africa during high-latitude glaciation was a two staged event with intermediate levels of aridity occurring from 50–32 ka cal yr BP followed by intense aridity from 32–18 ka cal yr BP.

The initiation of inflow from upstream Lake Kivu into Lake Tanganyika is evidenced at 10.6 ka cal yr BP through its influence on both elemental profiles (Mg, Ca) and through its effect on $^{87}Sr/^{86}Sr$. Increases in elemental (Mg, Ca, Sr) concentrations coincide with the timing of the Lake Kivu overflow. Metal geochemistry suggests that the overflow from Lake Kivu into Lake Tanganyika may have ceased between 8 and 6 ka cal yr BP, suggesting a period of Middle Holocene aridity in East Africa.

Keywords: Africa; Paleolimnology; Lake Tanganyika; Rift lakes; Late Pleistocene paleoclimate; Last Glacial Maximum

1. Introduction

Elemental geochemistry from sediment cores provides a powerful tool for reconstructing the paleoenvironmental and paleoclimate history of lakes and their watersheds (e.g. Mackereth, 1966, Ng and King, 2004; Roy et al., in

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press). Elemental geochemical methods have been used to infer changes in weathering intensity, water column and sedimentary redox conditions, and salinity in numerous paleolimnological investigations of north-temperate lakes (e.g. Mackereth, 1966; Engstrom and Wright, 1984; Davison, 1992). In contrast to temperate lakes and their watersheds, elemental geochemistry in tropical lacustrine sediment cores has been less frequently studied to infer weathering and environmental change (see Deevey et al., 1979; Curtis et al., 1998; Brown et al., 2000; Cardinal et al., 2001 for important exceptions). In this study we present sedimentary geochemical and sedimentological data from a 60,000-year-long core from Lake Tanganyika that demonstrates the value of elemental geochemical records for inferring paleoenvironmental variability within a tropical lake and its watershed. Our multi-indicator data provides new insight into the transition from humid to arid conditions experienced by central Africa during the northern hemisphere Last Glacial Maximum (LGM) and regional paleoenvironmental change over the past 60 ka.

Numerous studies have found that during periods of increased precipitation elemental concentrations of soluble elements such as K, Be, Mg, Ca, Ba and Sr, often increase in lake sediments, and conversely these same elements decline during dry periods (Mackereth, 1966; Engstrom and Wright, 1984; McHargue et al., 2000; McHargue and Donahue, 2005). These changes result from the fact that under conditions of rapid erosion, exposure and weathering of bedrock within a watershed, high concentrations of original silicate minerals enriched in soluble cations are eroded and transported to downstream lakes.

Elemental data from lake cores can be used to investigate paleolimnological processes as well. Fe and Mn are more soluble in anoxic conditions, thus low concentrations of these two elements have been interpreted to reflect sedimentary anoxia at the time of deposition (e.g. Haberyan and Hecky, 1987; Davison, 1992). As well the solubility and/or precipitation of numerous transition metals such as Ti, Co, Cu, Cr, Zn, V, Mn, Fe, Ni are sensitive to water column and sedimentary redox conditions (Mackereth, 1966; Davison, 1992; Brown et al., 2000). Concentrations of Co, Cr, Fe, Mn, Ni, Pb and Zn increase over an oxic–suboxic boundary (Balistieri et al., 1994), whereas there are noticeable decreases in concentrations of Cu and Cd under oxic conditions (Brown et al., 2000). Variations in lake level, windiness or lake temperature can all affect the relative position of the oxycline with respect to the lake floor, resulting in significant changes in redox-sensitive element concentrations. However, redox-sensitive elements must be used with caution to reconstruct anoxia, as their abundances can be affected by post-burial changes in redox conditions.

Changes in lake salinity also affect the elemental geochemistry of lake sediments. When salinity is high relatively insoluble metals are more likely to accumulate as precipitates in the sedimentary record, whereas during periods of dilute conditions, often accompanying high lake levels, these same metals are more likely to remain in solution (Davison, 1992). In addition, both elemental and isotopic profiles can be used to identify the timing of new hydrologic inputs (e.g. rivers, springs) into a lake, when those inflows carry unique or unambiguous chemical signals relative to the receiving basin (Hecky and Degens, 1973; Haberyan and Hecky, 1987).

Long sedimentary records from Lake Tanganyika indicate that, as in much of East Africa, aridity and lowered lake levels occurred during the Last Glacial Maximum (LGM) of higher latitudes (Gasse et al., 1989; Scholz et al., 2003; Talbot et al., 2006). Gasse et al. (1989) for example, suggested a drop in the level of Lake Tanganyika on the order of 400 m between 26.0 and 15 ka based upon fossil diatom assemblages in sediment cores from the southernmost part of Lake Tanganyika. Diatom and lithological data from those cores indicate the end of arid conditions and a major lake level transgression at about 15 ka BP, a signal confirmed by changes in the composition of organic matter (Talbot et al., 2006). The sequence of events prior to the LGM is much less clear. Scholz et al. (2003) argued, based on carbon isotopic data and the presence of a zone of woody debris in an offshore core from the Kavala Island Ridge, central Lake Tanganyika, for a shift towards conditions much more arid than today around 55 ^{14}C ka BP (57 ka BP) and a concomitant dramatic decrease in lake level. Fossil diatoms from this core show a similar sequence as the Gasse et al. (1989) record during and after the LGM, but are not preserved in sediments older than 28 ka BP.

Long paleoclimate records from East Africa are of importance for understanding climatic processes such as: (1) the role of insolation in regulating tropical climates at Milankovitch time scales; (2) the relationship between abrupt climate changes, variations in glacial ice extent, migration of the Intertropical Convergence Zone (ITCZ), and regional climate variability (Nicholson, 2000). Climate records from tropical Africa that extend beyond the last 25 ka are still quite rare, and long records from Lake Tanganyika are of particular interest, given its demonstrated potential for producing high-resolution (frequently annually laminated) sedimentary records (Cohen et al., 2006), and the ability to recover long, continuous, drill-core records of environmental change spanning millions of years. Here we present a new record from the central

basin of Lake Tanganyika that provides insight into the timing of critical climate events in East Africa during the Late Quaternary.

1.1. Lake Tanganyika: geography, limnology and regional climate

Lake Tanganyika, located between 9°S and 3°S, is the second largest freshwater lake in the world by volume (Fig. 1). It occupies a series of half graben basins in the western branch of the East African Rift Valley (Tiercelin and Mondeguer, 1991). The bedrock geology of the Lake Tanganyika basin is primarily Proterozoic metasedimentary rocks, with basaltic volcanic rocks occurring in significant quantities only in the upstream basin of Lake Kivu (Tiercelin and Mondeguer, 1991; Cohen et al., 2006). The lake is extremely deep (at >1400 m the second deepest in the world) and permanently stratified (meromictic). The lake is currently hydrologically open, but approximately 95% of its annual water income is lost through evaporation. Consequently, Lake Tanganyika is slightly saline (conductivity=670 μmho/cm; Cohen et al., 1997). Because the lake hovers near the hydrologically open/closed threshold, its level is extremely sensitive to watershed precipitation/evaporation ratios.

Lake level in Tanganyika has also been responsive to overflow events in upstream Lake Kivu, which lies in the western rift valley north of Tanganyika. Lake Kivu has overflowed intermittently through the Ruzizi River into Tanganyika starting at 10.6 ka BP, if not earlier

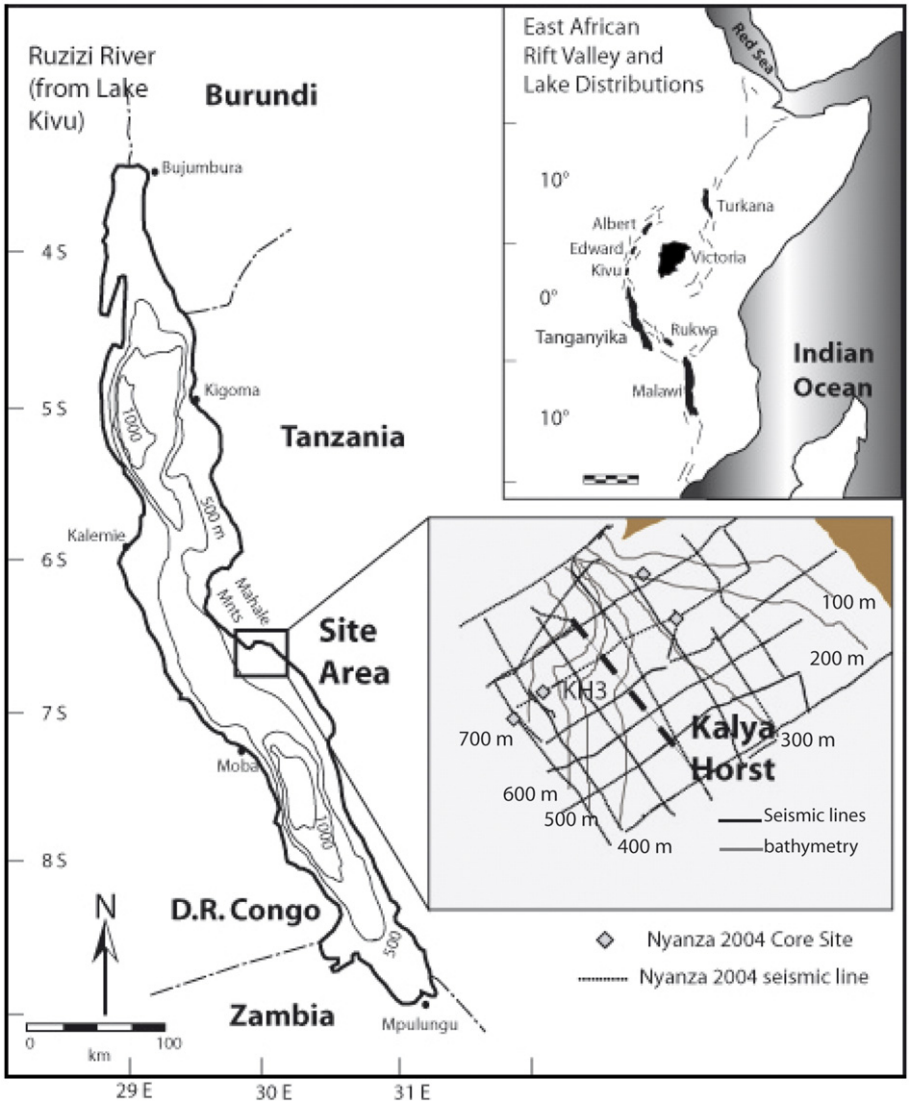

Fig. 1. Site map with core location and seismic lines from a survey conducted in 2004. Dashed line indicates general location of horst block.

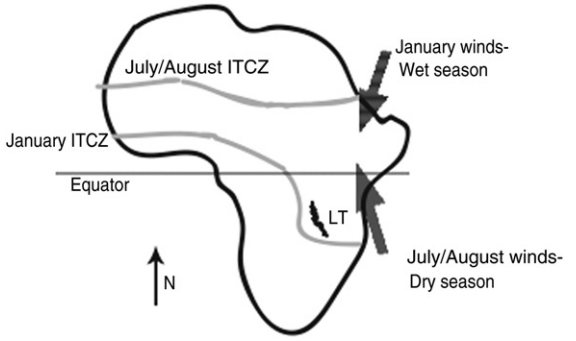

Fig. 2. Schematic of two different climate regimes over Lake Tanganyika (modified from Nicholson, 2000). Lake Tanganyika is denoted by (LT) and black shaded portion.

(Haberyan and Hecky, 1987; Stoffers and Hecky, 1978). Lake Kivu receives large hydrothermal inputs on the lake bottom; hence, Lake Kivu's chemistry (and the chemistry of the Ruzizi River) is more saline, enriched in numerous metals, and quite distinct from the chemistry (both elemental and isotopic) of Lake Tanganyika (Degens and Stoffers, 1976; Haberyan and Hecky, 1987; Barrat et al., 2000).

Lake Tanganyika lies within the migratory path of the ITCZ (Nicholson, 2000) (Fig. 2). The basin currently receives an average rainfall of 1200 mm/yr with a rainy season occurring between Sept. and May (Coulter and Spigel, 1991). The Mahale Mountains of the central Lake Tanganyika basin, adjacent to the core site discussed in this paper, have a significantly higher average rainfall (1800 mm/yr) than the lake basin average due to orographic effects (Cohen et al., 2006). Throughout the lake basin, a dry season between May and August coincides with strong winds from the south, whereas during the wet season winds are weaker (Plisnier et al., 1999).

Lake Tanganyika is permanently stratified, with anoxic bottom waters below ~100–130 m in our study area (Coulter and Spigel, 1991). Vertical mixing and partial ventilation of deep waters varies seasonally, as the thermocline tilts downwards towards the northern end of the lake during the dry windy season between May and September. This results in upwelling in the southern basin and the subsequent propagation of internal waves as the winds subside (Coulter and Spigel 1991; Plisnier et al., 1999; Naithani et al., 2003).

2. Methods

2.1. Coring location

In 2004 the Nyanza Project, an NSF-Research Experience for Undergraduates (REU) research training program on tropical lakes, collected a suite of piston cores from the Kalya horst block and platform, located at the north end of the southern basin of Lake Tanganyika (6°42.827′S, 29°49.957′E) (Fig. 1) using a modified Kullenberg piston corer (Kelts et al., 1986). The Kalya horst was chosen because gravity core and seismic stratigraphic studies of the area of the horst block indicated that this core site was in an area of continuous, but relatively slow sedimentation for at least the past 100 ka BP (Fig. 3). Seismic data show that our entire core record falls within a single stratigraphic sequence, and that a profound sequence boundary, indicative of considerably lower lake stands than any inferred from our record, underlies the base of our core by approximately 6.0 m (McGlue et al., 2006). Our most complete and temporally longest sediment core (NP04-KH3) was collected on the western side of Kalya horst in 640 m water depth. This 7.75 m core was shipped to the Limnological Research Center (U. Minn.) core lab, where it was split, digitally photographed and logged for magnetic susceptibility and gamma ray attenuation porosity (GRAPE) density using a GEOTEK core scanner.

2.2. Elemental and Sr isotopic analyses

The protocol for trace element digestion followed a procedure similar to that described by Hollocher et al. (1995). Approximately 0.08 g of dried sample powder was weighed into Parr® PTFE bomb liners. The samples were treated with 7 ml of a 9:1 solution of concentrated hydrofluoric acid (HF) and nitric acid (HNO_3), and evaporated to dryness on a hotplate and brought up into a solution of nitric acid. The sample containers were then sealed in steel bomb jackets and heated in an oven at 150 °C for 1 week. The samples were then removed from the oven and evaporated to dryness. The dried sample material was then treated with 2 ml of concentrated perchloric acid ($HClO_4$) and evaporated to dryness. Each of the samples were then treated with 15 ml of 6 M hydrochloric acid (HCl) and again evaporated to near dryness. The samples were redissolved in approximately 5 ml of 2 M HNO_3 and again heated in an oven overnight at 150 °C. The final solution was spiked with Re and In prior to analysis on a Perkin Elmer DRC II ICP-MS in the Department of Soil, Water, and Environmental Sciences at the University of Arizona. All acids used in the digestion process were distilled and when necessary diluted in high purity water in order to limit sample contamination. In order to quantify error associated with the analysis procedure, a lake sediment standard (LKSD-3, from Natural Resources of Canada) was analyzed

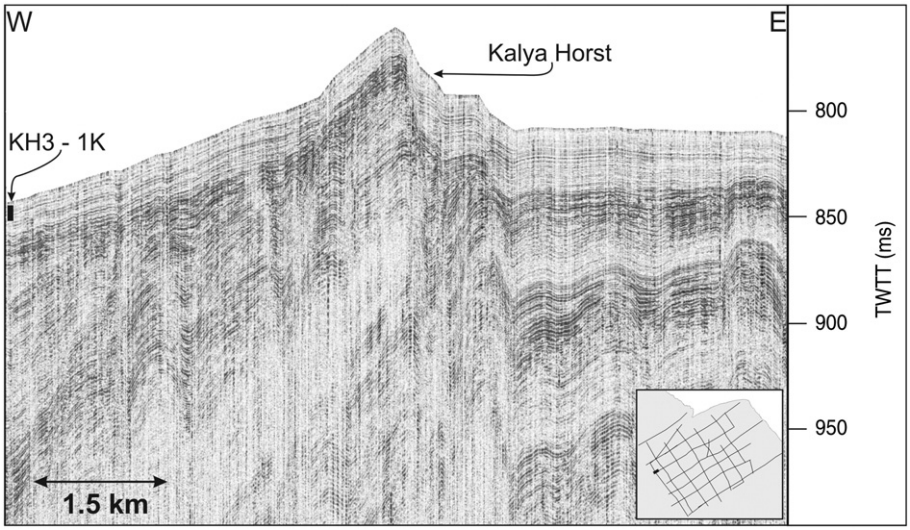

Fig. 3. Seismic line 6a adjacent to core NP04-KH3 location on the west side of the Kalya horst.

along with a sample blank for every 10 Lake Tanganyika samples. Twenty-two elements were analyzed, but three were below the detection limit of the ICP-MS. All elemental concentrations reported had an error less than 10% based on the stated value of the standard analyzed, sample blank, and instrument detection limits.

For Sr analysis, 20 to 100 mg samples were digested in HF and HNO_3 as described above. Strontium was separated utilizing Sr Spec resin (Eichrome Industries) using methods described in Brantley et al. (1998) and Chesley et al. (2002). Samples were dried and then loaded on tantalum filaments with Ta gel to enhance ionization and run on a VG sector 54 multi-collector thermal ionization mass spectrometer. Analyses of NBS-987 performed during the study, yielded a reproducibility of 0.710247 ± 0.000013 ($n=5$).

2.3. Biogenic silica and total organic carbon

Biogenic silica (BSi) and total organic carbon (TOC) measurements were made at the Limnological Research Center, University of Minnesota, Minneapolis.

Table 1
Radiocarbon age estimates (* denotes ages beyond Calib 5.0.1 range — see text for methods)

Lab accession number	Depth (cm)	Material	^{14}C age	Error (^{14}C yr)	Reservoir corrected (^{14}C yr)	Calib Age (yr BP)	1_Range	2_Range
WHOINOSAMS-50145	72.5	Bulk OM	4380	40	3970	4410	4410–4520	4300–4530
AA68366	129	Bulk OM	6690	45	6379	7300	7260–7410	7180–7420
AA61733	174	Bulk OM	7840	51	7560	8330	8340–8410	8210–8450
AA61732	174	Wood	8140	51	NA	8760	8590–8760	8550–8980
AA64614	210	Bulk OM	9110	60	8880	9970	9910–10,160	9740–10,190
AA67019	232	Bulk OM	9550	59	9330	10,500	10,430–10,650	10,300–10,700
AA64615	252	Bulk OM	10,270	62	10,080	11,660	11,410–11,810	11,340–11,970
AA67020	282	Bulk OM	10,620	58	10,440	12,380	12,170–12,590	12,110–12,640
AA67021	312	Bulk OM	12,170	84	12,060	13,920	13,820–14,000	13,750–14,100
AA61729	332	Wood	13,250	69	NA	15,930	15,500–15,890	15,340–16,110
AA61728	332	Bulk OM	13,320	71	13,250	15,720	15,500–15,890	15,330–16,110
WHOINOSAMS-50146	362	Bulk OM	14,650	70	NA	17,540	17,590–17,960	17,240–18,020
AA68364	382	Bulk OM	15,890	84	NA	19,100	18,970–19,160	18,920–19,300
WHOINOSAMS-50142	413	Bulk OM	24,600	170	NA	26,700	*	*
WHOINOSAMS-50143	473	Bulk OM	28,300	290	NA	31,900	*	*
AA68365	504	Bulk OM	30,780	390	NA	35,500	*	*
WHOINOSAMS-50392	563.5	Bulk OM	39,400	500	NA	41,800	*	*
WHOINOSAMS-50144	653	Bulk OM	44,500	760	NA	44,800	*	*
AA68966	593.5	Bulk OM	40,090	418	NA	42,000	*	*
AA68967	627.5	Bulk OM	38,480	384	NA	41,500	*	*

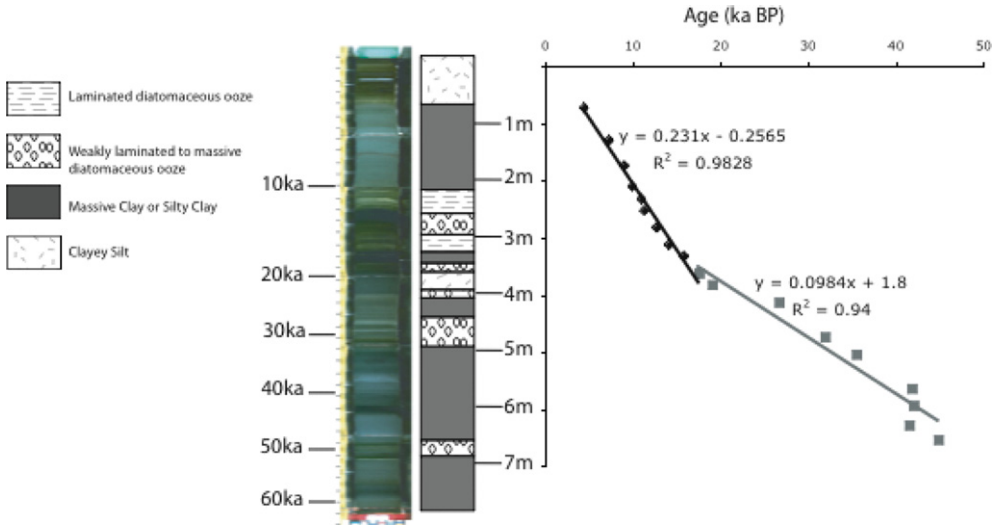

Fig. 4. Stratigraphy of sediment core NP04-KH3, and calendar year age model derived for this study.

Approximately 20 mg subsamples were analyzed for % BSi every 8 cm. The samples were analyzed for %BSi using multiple extractions of hot alkaline digestions at 85 °C in 0.5 M NaOH following the protocol of DeMaster (1979). BSi data were used to correct the total Si data from the elemental analyses above to calculate detrital Si. Total organic carbon (TOC) was measured using a UIC Inc. total carbon coulometer, correcting for carbonate concentrations using a UIC Inc. carbonate coulometer. This technique measures the amount of carbon dioxide during combustion of the sample with a reproducibility of ±0.2%.

2.4. Geochronology

AMS ^{14}C dating was performed on bulk sediment samples and wood samples at the Arizona-NSF Accelerator Mass Spectrometry Facility and the Woods Hole Oceanographic Institute National Ocean Sciences Accelerator Mass Spectrometry Facility. The chronology of the sediment core was determined from seventeen AMS measurements (Table 1; Fig. 4). The top of the core does not represent the most modern sediments due to over-penetration by the Kullenberg piston corer. The two oldest samples (40 and 38 ^{14}C ka) were prepared for AMS analysis in a new low background vacuum extraction system that is designed to provide reliable ^{14}C ages in the 40–60 ka range (Pigati et al., 2007).

Prior to calendar year calibration the raw radiocarbon ages were corrected for the reservoir effect of old carbon residing in the DIC pool of Lake Tanganyika (Fig. 5; Table 2). The reservoir effect was calculated from the offset in ^{14}C ages between paired bulk organic matter (sediment) and terrestrial plant material samples collected from the same stratigraphic horizons. The results of these paired radiocarbon age offsets appear to show a large

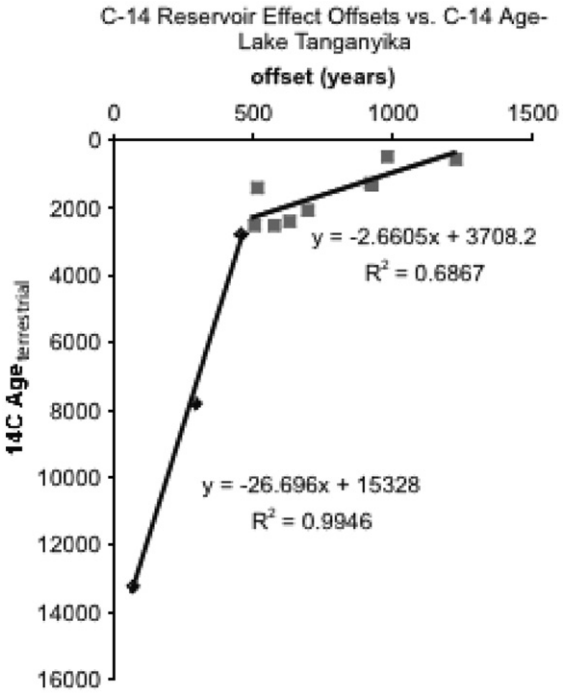

Fig. 5. Variability in old-carbon reservoir effect on bulk organic matter radiocarbon ages in Lake Tanganyika sediments. Offsets are expressed as ^{14}C yr$_{\text{bulk lake sediment organic matter}}$-^{14}C yr$_{\text{terrestrial organic matter from same horizon}}$. Data are from Cohen et al. (1997, 2006), Stager (unpublished) and this study.

Table 2
Radiocarbon reservoir effect (* denotes approximate location)

Location	Longitude	Latitude	Water depth (m.)	Age ^{14}C years	Offset ^{14}C years	1sd	Material
Luiche Platform, TZ LT-03-05	29.61E*	4.95S*	130	500	980	35	Stick
Luiche Platform, TZ LT-03-05	29.61E*	4.95S*	130	595	1225	35	Plant
Luiche Platform, TZ LT-03-08	29.60E*	4.97S*	336	1310	920	40	Wood
Kigoma Bay, TZ LT- 03-02	29.585E *	4.876S*	109	1360	925	35	Plant
Kigoma Bay, TZ LT-03-02	29.585E *	4.876S*	109	1435	515	35	Plant
Kigoma Bay, TZ LT-03-02	29.585E *	4.876S*	109	2100	695	35	Plant
Luiche Platform, TZ LT-03-05	29.61E *	4.95S*	130	2430	670	35	Charcoal
Kigoma Bay, TZ LT-03-02	29.585E *	4.876S*	109	2550	505	35	Wood
Kigoma Bay, TZ LT-03-02	29.585E *	4.876S*	109	2480	575	40	Wood
Kalya Platform Slope, TZ (LT-00-02)	29 58.306E	6 33.154S	309	2845	458	48	Wood
Kalya Platform Slope, TZ (NP04-KH1)	29 58.480E	6 33.147S	303	7838	297	51	Wood
Kalya Platform Slope, TZ (NP04-KH1)	29 58.480E	6 33.147S	303	13,250	71	69	Wood

increase in the reservoir of old carbon in Lake Tanganyika during the late Holocene relative to the late Pleistocene/early Holocene. The regression-defined age offset declines to zero for sediments older than 14 ka ^{14}C, and no corrections were applied to age dates older than this. The origin of this change is unknown, but we speculate

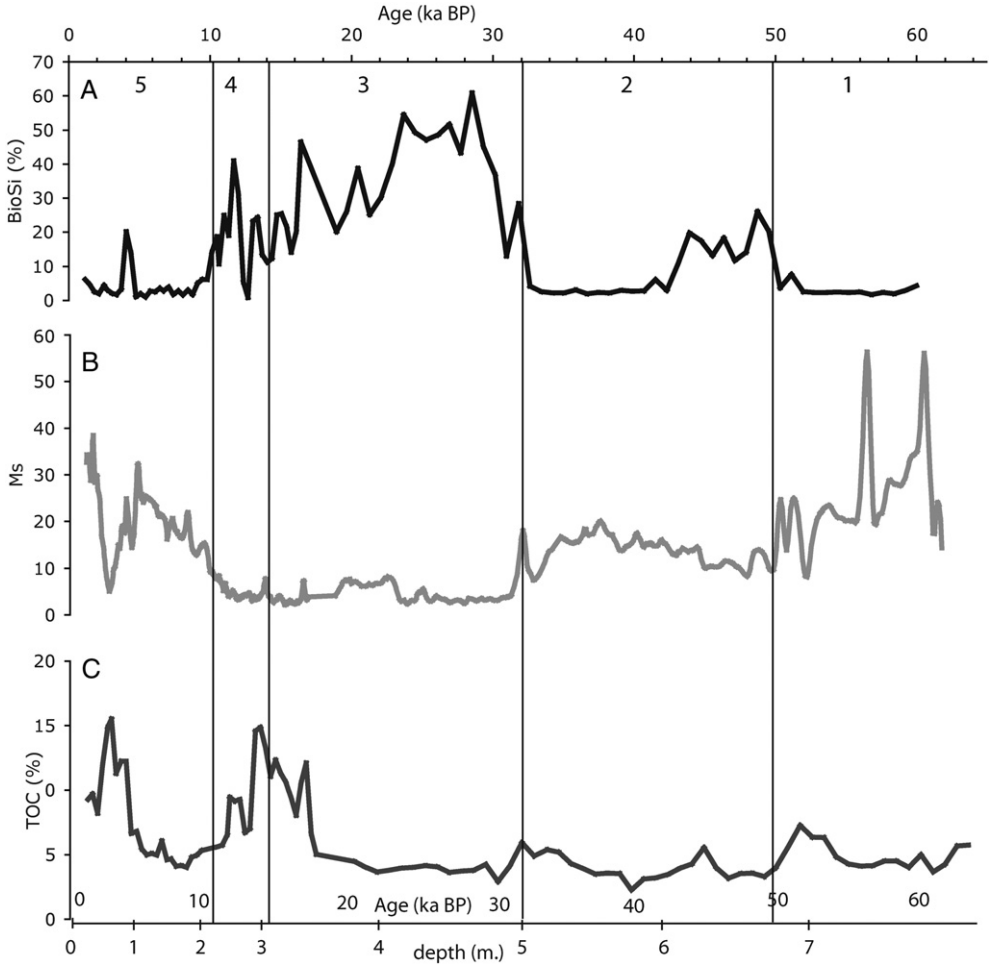

Fig. 6. Physical stratigraphy of NP04-KH3 cast against calendar age for: A) biogenic silica, B) magnetic susceptibility, C) total organic carbon. The lines indicate boundaries of climatically-linked stratigraphic zones discussed in the text.

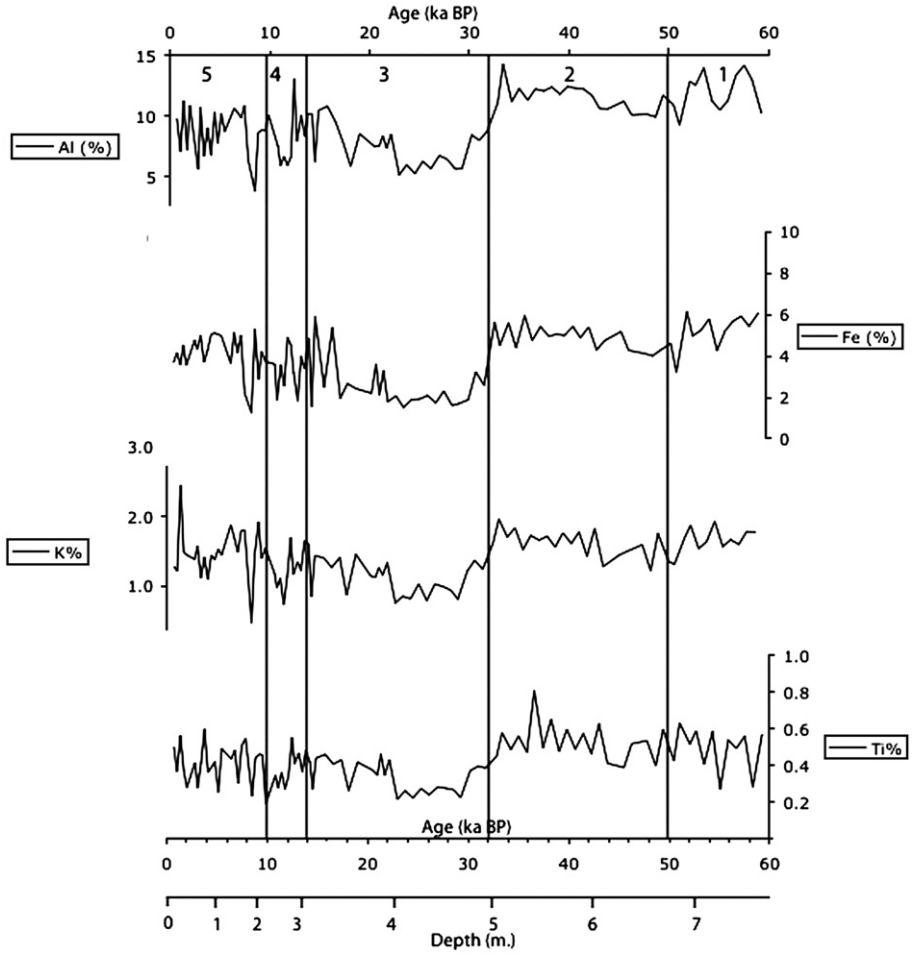

Fig. 7. Concentrations of Al, Fe, K and Ti in the NP04-KH3 core record. The lines indicate zonal boundaries discussed in the text.

that enhanced water column stratification, combined with longer water residence times during the past 2000 yr compared to the early Holocene, have allowed for old carbon to accumulate in the lake. Because only the older three paired sample age offsets (which define a steeper slope on Fig. 5) are from the immediate area of this study, it is possible that there have been different residence times in the Kalya region as compared to other parts of the lake. However, the effects of variations in this offset history are minimal for our record. This is because (1) our regression implies no apparent offset in the portion of the record older than 14 ^{14}C ka, which makes up the bulk of our history, and (2) only a few data points from the record presented here are young enough at the top of the core to have significant age offsets.

Corrected radiocarbon age estimates younger than 26 ka ^{14}C were calibrated with Calib 5.0.1 (Stuiver and Reimer, 1993; Reimer et al., 2004), and the older radiocarbon age estimates were calibrated using data from Hughen et al. (2004). Two linear regressions were used to accommodate varying sedimentation rates throughout the sedimentary record (8.5 cm kyr^{-1} from 7.75 m to 3.30 m vs. 22.4 cm kyr^{-1} from 3.30 m to 0 m). Age estimates for the bottom sections of the record (below the lowest ^{14}C date) were estimated by extrapolation of the lower core section's age model. All age estimates discussed in this paper are expressed in calendar years BP unless otherwise indicated.

2.5. Statistical analyses

All time series data was plotted in Excel®. A principal components analysis was conducted to understand the relationship of elemental chemistry and physical properties data within and between samples in the sedimentary record. Principal components analyses on elemental ratio

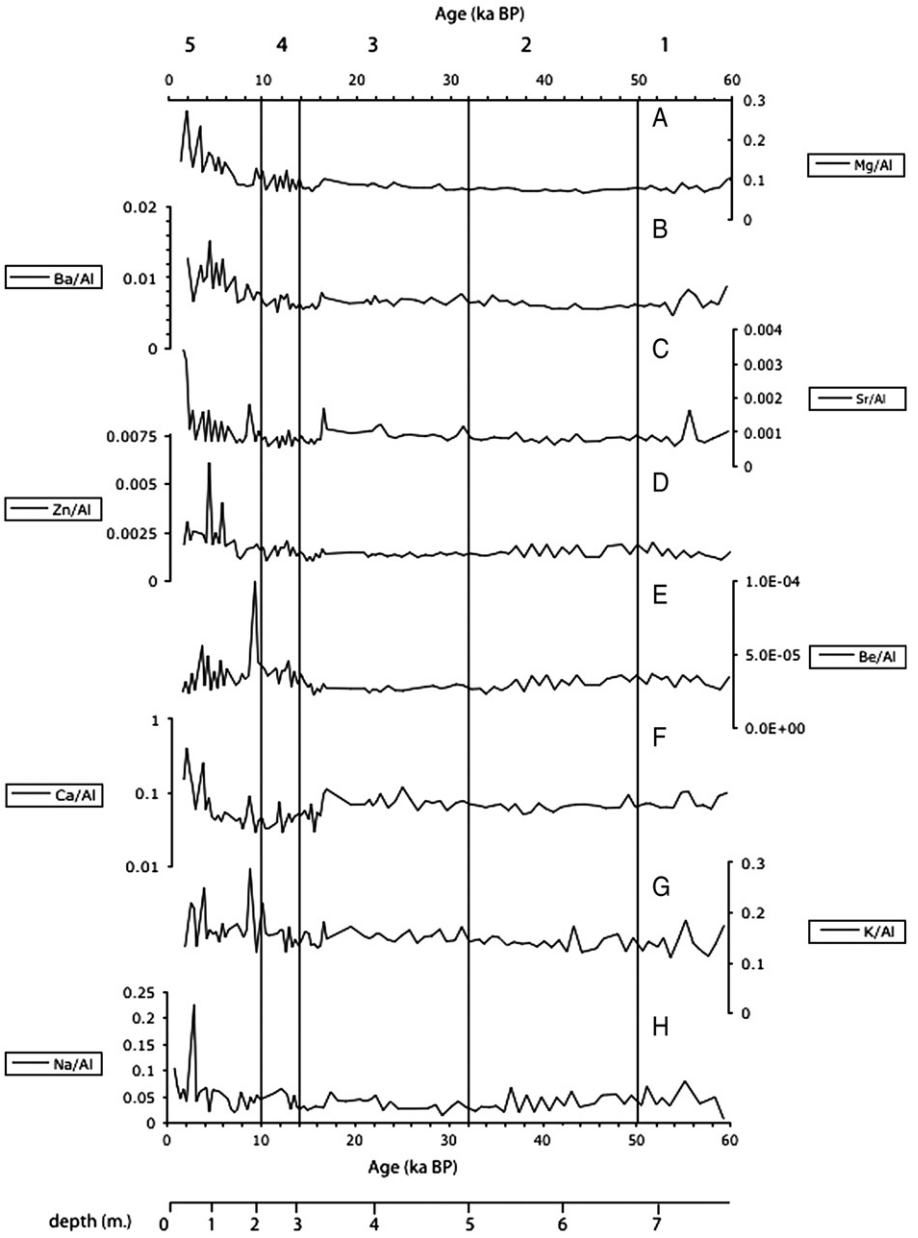

Fig. 8. Elemental ratios relative to Al for A) K, B) Ba, C) Sr, D) Zn, E) Be, F) Ca, G) K, H) Na cast against age and depth in NP04-KH3. The lines indicate zonal boundaries discussed in the text.

data for all elements above detection limits were performed with JMP IN 5.1 for Macintosh.

3. Results

3.1. Lithostratigraphy, physical properties, biogenic silica and TOC

NP04-KH3 is characterized by primarily massive, light to dark gray silty clay with occasional diatomaceous beds from its base at 7.75 m, (~60.2 ka) to 3.38 m (15.6 ka) (Fig. 4). Within this interval, from 4.98 m (32.3 ka) to 3.57 m (16.5 ka) the core is characterized by weakly bioturbated sediments. The overlying sediments between 3.38 m (15.6 ka) to 2.25 m (10.9 ka) consist of laminated diatomaceous ooze alternating with dark organic rich horizons. From 2.25 m to 0.80 (10.9 ka to 4.7 ka) the core consists of massive clay, and the uppermost portion of the core (0.80–0.00 m) is very finely laminated silty clay.

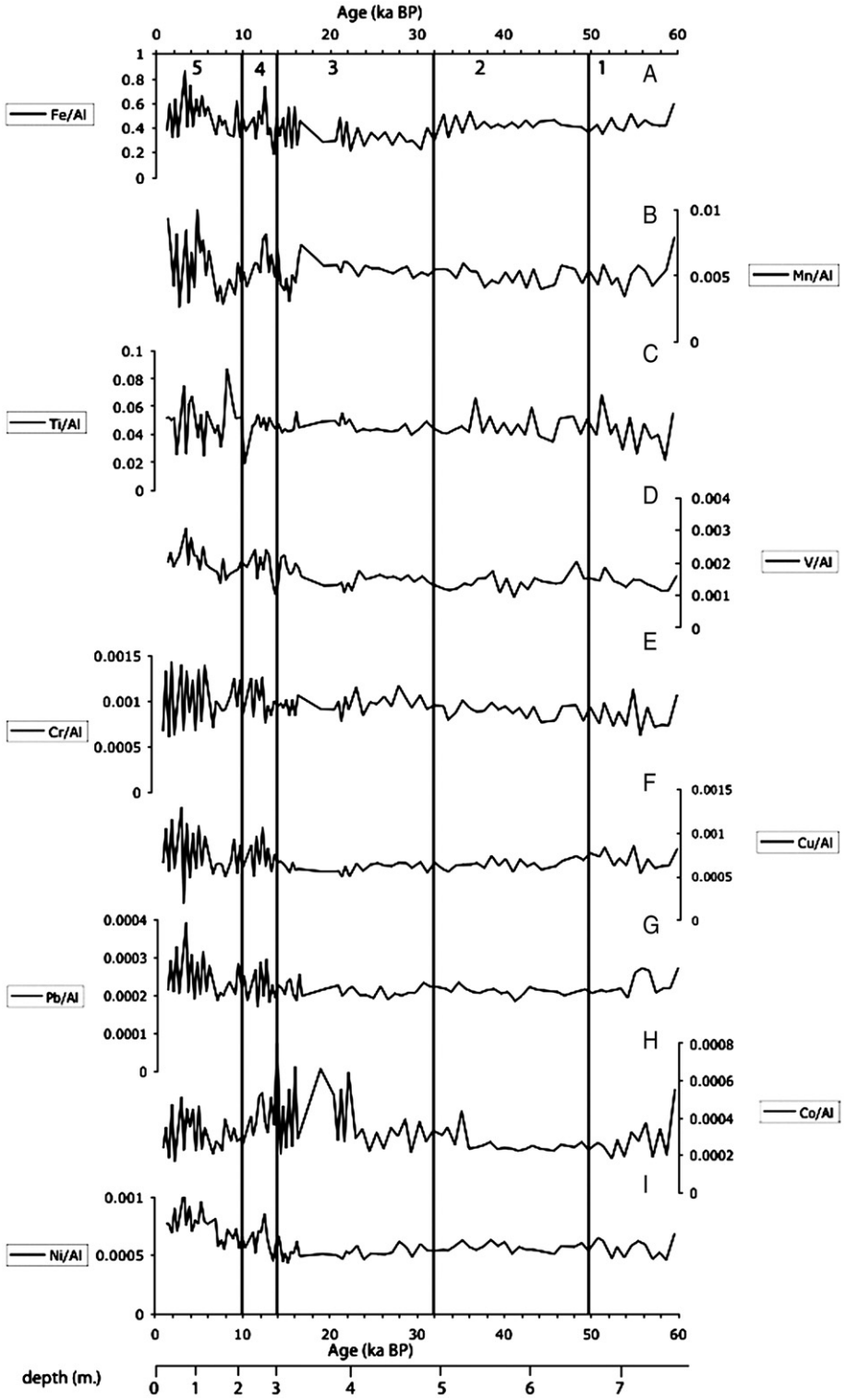

Fig. 9. Elemental ratios for the NP04-KH3 core ratioed against Al for; A) Fe, B). Mn, C) Ti, D) V, E) Cr, F) Cu, G) Pb, H) Co, and I) Ni. Bars indicate zonal boundaries discussed in the text.

The biogenic silica concentration remains low (<20%) throughout the record except between 4.98 and 3.59 m (32.3–16.6 ka) where the biogenic silica increases (38.5±12.9%). The magnetic susceptibility record is characterized by moderate to high values from the base of the core to 5.07 m (33 ka) (17.6±8.7) (Fig. 6). The middle section 5.07–2.20 m (33–10.4 ka) of the record is characterized by low to moderate levels (4.0±1.8), and the upper portion of the record has increased values (17.6±8.0). Total organic carbon throughout the record remains relatively low to moderate throughout the lower portion of the sedimentary record up to 3.59 m. From 3.59 to 3.02 m (16.6 to 14.1 ka) there are increasing values of TOC (9.6±1.9%). TOC is low between 1.67 and 0.62 m (8.4 to 3.9 ka) (4.9±0.9%), but increases dramatically at the top of the sedimentary record (11.5± 2.4%).

3.2. Elemental analyses

Trends in Al, Fe, K and Ti (as well as many other elements) show a strong inverse correlation with biogenic silica because the latter strongly influences the bulk composition of the core (Fig. 7). Therefore, we will present all subsequent discussion of our elemental data in terms of ratios against Al, the most insoluble (under both oxic and anoxic conditions) and common, terrestrially-derived fraction (Brown et al., 2000) (Figs. 8 and 9).

Between 60 and 32 ka (7.75 to 4.98 m) values of Fe, Be, Na (and more variably Ti) ratios against Al are higher than in the following interval. Ratios of these elements, as well as Zn, Mn, K) are also more variable from 60–32 ka. From 32 ka–18 ka BP (4.98–3.59 m) Zn, Na, Be, Fe, Cu, and possibly Ti to Al ratios decline. Other elements aside from Co show no significant trend in this interval relative to the earlier one. Co/Al begins to increase at ~32 ka BP and rises dramatically after 22–10 ka BP (Fig. 9). At around 18 ka Fe and V begin to increase, whereas Mn, Ti, Cr, Cu, Pb, K, Ni and Be ratios begin to increase later, between 14.3 ka BP (e.g. Mn) to around 12 ka BP, but with the greatest change in all these elements after 12 ka BP (Figs. 8 and 9). Mg, Ba, Sr, Co and Zn increase dramatically at 8 ka BP (Fig. 8). The ratios of Na/Al and Ca/Al increase significantly in the last 0.8 m of the sedimentary record at 4 ka BP.

Several elemental ratios (Fe, Mn, Ti, V, Cr, Cu, and Pb) increase in two distinct periods over the Holocene (Fig. 8). These elemental ratios are elevated between 12 and 8 ka BP (2.60–1.88 m) and again between 6 ka BP (1.18 m) and the top of the sedimentary record.

3.3. Principal component analyses

The principal components analysis of elemental concentration/Al concentration ratio using correlation data indicates two major components of variance in the data (Fig. 10; Table 3). The first component accounts for 35.55% of the variance (eigenvalue=6.4), and shows a strong increase at the Pleistocene/Holocene transition. Elemental ratios (to Al) with strong loadings on the first PCA axis include Mg, Ni, Pb, Cu, Ba, Mn, Cr, K, and V. The first PCA axis appears to combine elements with relatively low variability in the Pleistocene with strongly rising concentrations in the Holocene.

The second PCA axis for the elemental ratio data accounts for 13.3% of the total variance (eigenvalue=2.4), and corresponds to the dramatic decrease in total soluble cation concentrations between ~30 and 17 ka BP. Elemental ratios (to Al) with strong positive loadings on the second PCA axis are Sr, Ca and K,

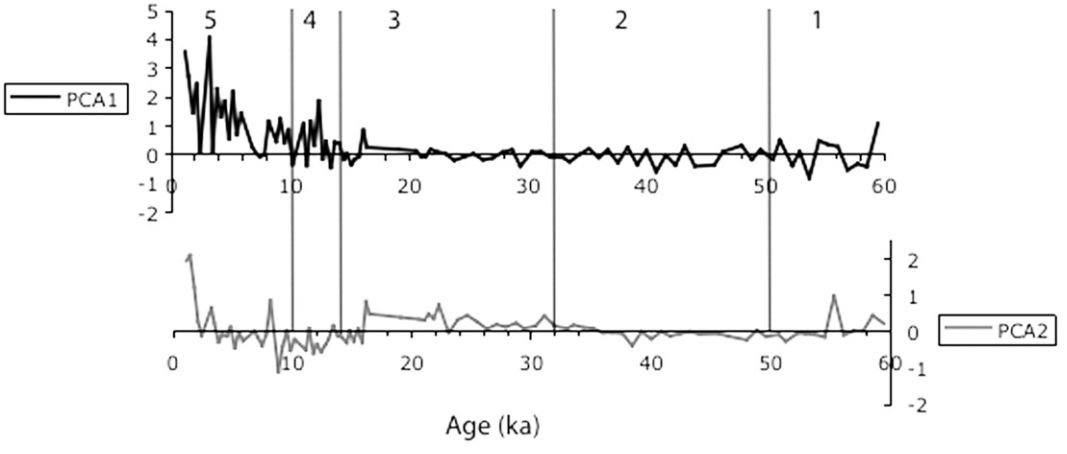

Fig. 10. Principal components axes (PCA) 1 and 2 from metal/Al ratio and (correlation matrix) plotted against time.

Table 3
Principal components analysis (PCA) of Me/Al data from NP04-KH3. Moderately–strongly positive and negative loading ratios are underlined

Principal components: on correlations		
Eigenvalue	6.393	2.4362
Percent	35.5165	13.5343
Cum percent	35.5165	49.0509
Eigenvectors		
Fe/Al	0.14918	−0.24528
Mg/Al	0.31966	0.02641
Ca/Al	0.06388	0.45233
K/Al	0.25409	0.29563
Na/Al	0.14941	0.12725
Ti/Al	0.23735	−0.04302
Ba/Al	0.26523	0.18228
Zn/Al	0.22461	−0.13511
Mn/Al	0.2592	0.19789
Sr/Al	0.2178	0.45161
V/Al	0.25124	−0.2097
Ni/Al	0.31619	−0.13395
Cr/Al	0.25544	−0.17364
Cu/Al	0.26941	−0.30953
Co/Al	0.17221	0.20454
Pb/Al	0.28586	0.07268
Be/Al	0.2035	−0.28104
Tl/Al	0.19778	−0.14328

whereas there are strong negative loadings for Cu, Be and Fe. The contrast in loadings might be expected if the axis is driven to higher values by an overall decrease in precipitation, which would lead to less Be weathering and increased Ca and Sr precipitation. However, the lack of TIC from 30–17 ka BP indicates little Ca and Sr precipitation, and instead suggests the influence of Lake Kivu input. The strong influence of Lake Kivu tends to mask the more subtle record of the metals earlier in the sedimentary record.

3.4. Sr isotopes

The $^{87}Sr/^{86}Sr$ values from core samples between 23.8 and 9.3 ka BP are very similar and range between 0.738 and 0.740. However during the Holocene the values decrease markedly to 0.725 at 8.4 ka and appear to recover to more radiogenic pre-Holocene values at 8 ka and then gradual decrease to $^{87}Sr/^{86}Sr$ values of ~0.725 at 1.6 ka BP (Fig. 11).

4. Discussion

The geochemical record from core NP04-KH3 shows considerable variability over the past ~60 ka BP, which can be interpreted in terms of variation in weathering rates, variable redox conditions in bottom waters, and the influence of upstream contributions from Lake Kivu (Fig. 12).

4.1. 60 to 50 ka BP — zone 1

The relatively high but variable magnetic susceptibilities, high Fe/Al, and moderately high Be/Al ratios of this interval all suggest relatively strong terrestrial weathering intensity and probably high precipitation/evaporation (*P*/*E*) ratios within the adjacent Mahale Mountains watershed. Similarly, low relative concentrations of the divalent cations (Ca, Mg and Sr) point towards little or no carbonate precipitation, also consistent with low salinities and relatively humid conditions during this time. There is evidence of strong weathering of feldspars characterized by increased values of K/Al throughout this interval.

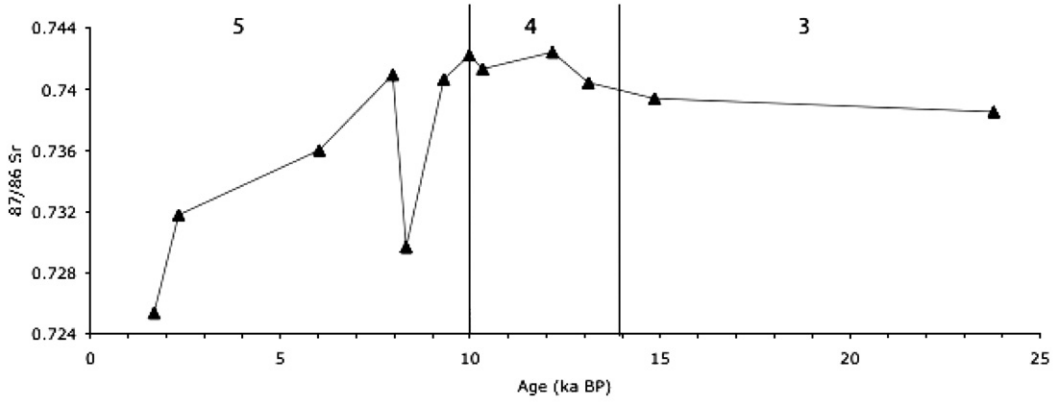

Fig. 11. Strontium isotope values prior and after the Lake Kivu overflow with zonal boundaries.

4.2. 50 to 32 ka BP — zone 2

Decreased magnetic susceptibility, a shift towards lower and less variable elemental ratios for a number of elements (notably in Co and Fe), a shift in lithology to massive, silty clays in the older portion of the interval (50–43 ka BP) and diffuse laminations later (43–32 ka BP), all suggest a paleoenvironmental transition at 50 ka in the Tanganyika basin. This transition is also marked by slightly elevated Ti/Al, V/Al and Cu/Al between 50 and 43 ka BP and a return to lower levels of all of these from 43–32 ka BP. The reduced values of magnetic susceptibility could be caused by a reduction in watershed weathering rates and delivery of suspended terrigenous sediment to this topographic high, suggesting a shift toward reduced weathering intensity and a somewhat lower *P/E* ratio in the surrounding watershed during the 50–32 ka BP interval relative to the preceding period. Increases in the depth or stability of the mixing zone and oxycline between 50 and 43 ka BP relative to the preceding time interval and the 43–32 ka BP period are indicated by elevated ratios of Cu/Al.

4.3. 32 to 14 ka BP — zone 3

This section of the sedimentary record is characterized by bioturbated sediments, indicative of significantly lowered lake levels or deeper water column mixing induced by stronger winds and lower temperatures. Low TOC and low magnetic susceptibility shifts are both consistent with relatively drier conditions during this interval.

4.4. 14 to 10 ka BP — zone 4

Increases in Be/Al, Ni/Al, Cr/Al, Fe/Al indicate a transition from the previous drier period to a wetter interval with increased weathering intensity in the surrounding watershed. Although these trends are clear, the beginning of this transition is difficult to pinpoint as different indicators suggest slightly different timings and rates for this transition. Elemental indicators of increased watershed weathering suggest an arid/humid transition between about 18 and 15 ka BP, coincident with the transition from diffuse to finely laminated sediments and shifts in TOC and BSi. The second PCA axis (Fig. 10) shows increases in elemental concentrations throughout this section that are also consistent with rising lake level conditions.

Between 12 and 10 ka divalent cation ratios are low, redox-sensitive elemental abundances (Cr, Cu, Co, Mn, and V) are high, and Fe remains variable, suggesting a short-lived deepening in the oxycline during the arid/humid transition.

4.5. 10 ka to 0.8 ka BP — zone 5

Changes in metal/Al ratio trends can be explained by increased weathering intensity in the Lake Tanganyika watershed under the more humid climate of the early Holocene as compared with older sediments. However, these trends must also reflect a new source of metals to Lake Tanganyika during the Holocene relative to the Pleistocene portion of our record. Almost certainly these trends record the first influxes of relatively hard and saline Lake Kivu waters (Haberyan and Hecky, 1987). Increased soluble cation ratios during the Holocene coincide with the dramatic shift in $^{87}Sr/^{86}Sr$ isotopic composition of lake sediments in the early Holocene (~8 ka), interpretable as a signal of the influx of Tertiary basalt-derived unradiogenic Sr from the Kivu basin into Lake Tanganyika. Previously Sr input into Lake Tanganyika was derived exclusively from older more radiogenic Precambrian and Karoo, Late Paleozoic-aged sources (Fig. 11). Some hydrothermal activity has been recorded in the northern portion of Lake Tanganyika (Tiercelin and Mondeguer, 1991), and it might be argued that the dramatic shift in Sr isotopes is a record of hydrothermal activity changes within Lake Tanganyika independent of Kivu sources. However Barrat et al. (2000) analyzed the strontium isotopes in gastropod shells and lake water from northern Lake Tanganyika near hydrothermal vents. Their results showed isotopic ratios of $^{87}Sr/^{86}Sr$ from 0.7152 to 0.7165 for lake waters and 0.72183 to 0.72495 for the hydrothermal aragonite chimneys. Thus the primary source of the $^{87}Sr/^{86}Sr$ ratios in hydrothermal inputs within Lake Tanganyika proper is also a reflection of older, more radiogenic sources and not Tertiary magmatic sources. Our $^{87}Sr/^{86}Sr$ ratio differences prior to the Holocene (0.738 and 0.739) vs. during the Holocene (0.731 and 0.725) therefore provide an unambiguous indication of either continuous or episodic overflow of Lake Kivu into Lake Tanganyika.

Hard and saline Kivu water input may be accentuated in the Late Holocene, consistent with earlier investigations of carbonate accumulation and saturation in Lake Tanganyika (Cohen et al., 1997; Alin and Cohen, 2003). There is no carbonate in Lake Tanganyika sediments prior to ~3 ka BP, but authigenic, inorganic calcite and aragonite are present throughout the lake in sediment younger than ~3 ka BP, including our sediment core. Apparently even during the relatively arid conditions of the Late Pleistocene the solute load derived from relatively insoluble bedrock sources surrounding Lake

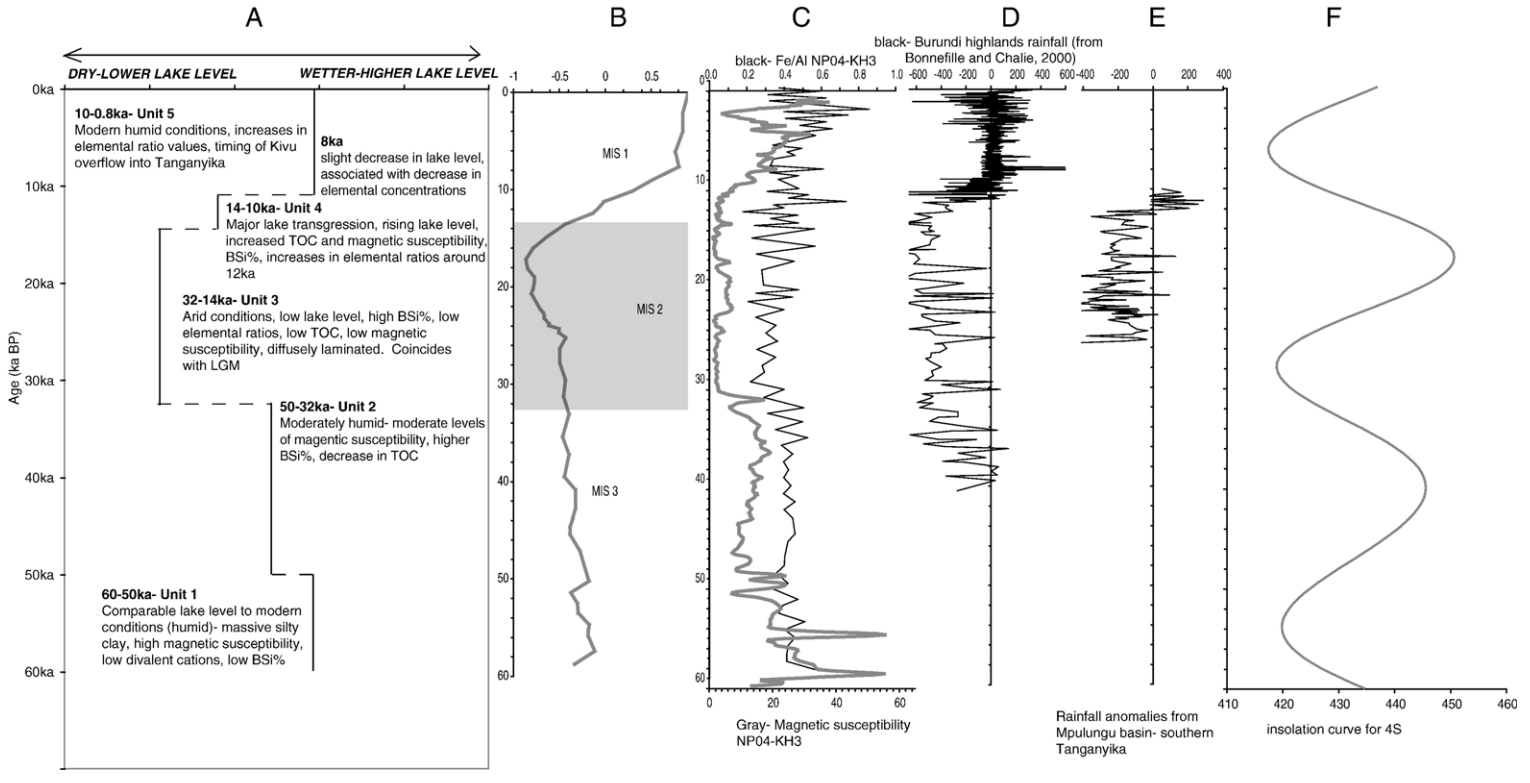

Fig. 12. Summary figure of climate and lake level fluctuations in Lake Tanganyika over the past 60 ka, B) $\delta^{18}O$ from marine record Martinson et al. (1987); C) Fe/Al (black) and magnetic susceptibility (gray) for NP04-KH3; D) estimation of rainfall as it deviates from modern day mean annual rainfall from Burundi highlands (black) from (Bonnefille and Chalie, 2000); E) estimation of rainfall as it deviates from modern day mean annual rainfall from Mpulungu basin (southern Tanganyika) (Bonnefille and Chalie, 2000); F) insolation at 4°S.

Tanganyika was insufficient to raise the lake's chemical concentration to the point of carbonate saturation, whereas this could occur under much less arid conditions during the Holocene with the addition of Kivu's saline hydrothermal waters. Based on elemental and isotopic data it does not appear that Lake Kivu overflowed into Lake Tanganyika until the early Holocene, following the ~14–11 ^{14}C ka BP blockage of Lake Kivu's outlet to the north by the Virunga volcanics (Coulter and Spigel, 1991). The significant decrease in the strontium isotope values highlights the input of elemental rich waters from Lake Kivu entering Lake Tanganyika during this interval. This would explain why similar weathering intensities between the 60 and 50 ka BP period and the late Holocene (reflected in magnetic susceptibility trend) are marked by such different metal ratios.

The increase in concentration of soluble cations during the Holocene appears to have occurred in two phases (Fig. 9), interrupted by an interval of lower metal/Al ratios that may indicate a brief cessation between ~8 and 6 ka in the concentrated deposition of Kivu-derived metals. At that time Lake Kivu may have ceased flowing causing soluble metal ratios to decrease, consistent with some other Middle Holocene records that suggest relative aridity at that time (Thevenon et al., 2002). Given uncertainties in dating this period may also coincide with pronounced aridity in northern Africa as recorded in Lake Bahar-El-Ghazal and Lake Abhe between 9 and 8 ka (Gasse, 2000). The increase in all the elements in the late Holocene (~4 ka BP to top of core) is consistent with a renewal of Lake Kivu overflow into the Tanganyika basin in the Late Holocene.

4.6. Correlation of NP04-KH3 with other regional and global paleoclimate records

Several paleoclimate records are available for East Africa that cover some or the entire interval recorded in the NP04-KH3, allowing us to evaluate our record within a regional context. Our inference of aridity in the late Pleistocene followed by a dramatic lake level rise and wetter climates during the latest Pleistocene is consistent with a variety of studies from both Lake Tanganyika (Haberyan and Hecky, 1987; Gasse et al., 1989; Talbot et al., 2006) and, in fact, most of East Africa (Gasse, 2000). However, the NP04-KH3 record provides a number of important new insights into regional paleoclimate, in some cases clarifying the timing of important climate transitions, and in others apparently contradicting earlier inferences.

Our record shows evidence for relatively moist conditions between 60 and 50 ka BP, roughly comparable to modern conditions, followed by a transitional period of moderately humid conditions from about 50–32 ka BP and a sharp onset of aridity thereafter. The only other published record from the southern tropical African continent that extends back to 60 ka is that of Scholz et al. (2003) from the Kavala Ridge of central Lake Tanganyika. Based on TOC, lithologic, and stable isotope data those authors argued for an arid period starting around 57 ka BP and continuing to the latest Pleistocene, with brief pulses of stronger aridity at 42, 29 and 23 ka ^{14}C BP (44, 35, 26 ka cal yr, respectively). Their interpretation of the early (57 ka) onset of aridity was based primarily on a shift toward increased δ^{13}C, which they interpret to reflect increased inputs of C_4 grasses to the lake. However, δ^{13}C values in organic matter are sensitive to a variety of terrestrial and limnological processes, and Talbot et al. (2006) indicate that the organic matter and the δ^{13}C signal in Lake Tanganyika is primarily lacustrine and may not reflect terrestrial inputs. A thin bed of woody material in the Kavala sediment core, which Scholz et al. (2003) suggest marks a major (ca. 300–350 m) lowstand at about 57 ka BP is not apparent in our data. Scholz et al. (2003) note that the woody material could be derived from floating vegetation rather than in situ aquatic macrophyte/plant growth, the latter formed during lowstand conditions. The transition at 50 ka BP in the Kalya record may be related to the change at 57 ka (55 ^{14}C ka BP) in the Kavala site, as neither transition is directly dated. Secondly, our data do not confirm the 350 m lowstand inferred by Scholz et al. (2003) at this time.

A major result of our study is a clear indication of the onset of increased aridity in the Lake Tanganyika basin associated with the onset of higher latitude glaciation (MIS 2) at 32 ka BP. Estimates of the timing of the onset of aridity associated with MIS 2 in tropical Africa vary widely: 23 ka BP at Lake Albert (Beuning et al., 1997), 35 ka BP at Lake Turkana (Johnson, 1996), 22 ka BP at Lake Abhé (Gasse, 1997), 22 ka BP at Lake Tritrivakely (Gasse and Van Campo, 2001), and 35 ka BP in Burundi (Bonnefille and Chalié, 2000). Some of these differences may reflect real, regional variability in climate conditions in East Africa, others may reflect difficulties in dating due to discontinuous sedimentation. Our finding for a humid–arid transition at 32 ka BP is in general agreement with pollen stratigraphic evidence in the equatorial highlands of East Africa that conclude that arid conditions prevailed generally in East Africa from 35–18.3 ka BP (30 to 15 ka ^{14}C BP; Bonnefille and Chalie, 2000; Fig. 12), suggesting at least similar timings for this part of tropical Africa.

Gasse et al. (1989) argued for an onset of arid conditions in Lake Tanganyika at 26 ka BP based on an

abrupt increase in benthic/planktonic diatom ratios in cores from the southern basin of the lake. We suggest that this timing (considerably later than our proposed onset of aridity) is a consequence of a taphonomic artifact of diatom accumulation in deep water environments. Because this rift lake has extremely steep lake floor slopes in its upper few hundred meters and much gentler slopes in deeper water, the proportion of littoral habitat (where benthic diatoms can be produced) to deepwater environments is relatively insensitive to lake level declines in Tanganyika of up to several hundred meters. Such proportional differences in lake floor morphometry are known to have significant effects on the accumulation of benthic diatoms (Stone and Fritz, 2004). Furthermore, once falling lake level has declined sufficiently to expose the more gently sloping, previously deep water environments, erosion of the now more-expansive littoral deposits would yield an even higher proportion of benthic diatoms to offshore deep zones. This combination of effects could produce a benthic/plantkonic diatom ratio record that is both insensitive to the early effects of lake level fall, and then magnifies the ratio signal once the lake falls below the morphometric threshold of decreasing bottom slope. However, the timing of the transition to higher lake levels would be recorded accurately by benthic/planktonic fossil ratios. This is significant as all records from Lake Tanganyika and most from other regions of tropical Africa, whether derived from terrestrial or aquatic indicators, are in agreement on the timing of the arid/humid deglacial transition.

Gasse et al. (1989) found evidence for the lowest lake levels in Tanganyika about 21.2 ka BP, consistent with our study and the record from Kavala Island Ridge (Scholz et al., 2003). Evidence from Lake Tritivakely in Madagascar indicates an onset of precipitation at 17 ka BP (Gasse and Van Campo, 1998). Talbot et al. (2006) suggested that a major transgression in the Mpulungu basin of Lake Tanganyika began between 20 and 18 ka BP. Based on the diatom record from Lake Massoko, a small volcanic crater lake located between Malawi and Tanganyika, Barker and Gasse (2003) proposed that arid conditions peaked between 22 and 17.5 ka BP. Stager et al. (2002) analyzed diatoms from Lake Victoria and indicate that the lake level dropped dramatically between 18 and 17 ka BP and also between 15.9 and 14.2 ka BP. Beuning et al. (1997) showed that in the Lake Albert region of the western rift valley, north of our study area, aridity persisted until 14.6 ka BP. Elsewhere in the region, Filipi and Talbot (2005) inferred from Lake Malawi records that the post-LGM trangression began at 17.9 until 16.5 ka BP. In contrast, Johnson et al. (2002), in their analysis of the transition of periphytic to planktonic diatom dominance in the same lake, suggested that lake levels began to rise at 15.7 ka BP. Street-Perrott et al. (2004) examined a sediment core from Sacred Lake on Mt Kenya and suggested the onset of wetter conditions at 14.3 ka BP. Records from the northern subequatorial region of the Indian Ocean (Gulf of Aden, Arabian Sea) recording conditions in the Sahara and Arabian Peninsula were used to suggest an onset of wetter conditions at 14.8 ka BP, based on dust accumulation (deMenocal et al., 2000), or 15 ka BP based on trace element geochemistry (Sirocko et al., 2000), both of which post-date the timing of evidence from the Kalya Horst region for increased weathering and rising lake levels.

Thus, a picture emerges in which extreme aridity coincident with the LGM ended by about 14.5 ka BP, often preceded by more subtle shifts between 22 and 18 ka, although many regional records do not have continuous sedimentary sequences that extend beyond ~18 ka BP, limiting inferences regarding the timing of climate changes prior to the end of African aridity at ~14.5 ka BP. Our record, based on both sedimentology and elemental data, indicates a gradual transition to more humid conditions during the late Pleistocene, such as increases in metals/Al, do not show clear changes until ~15 ka BP. Thus, our data also suggests an early, subtle shift toward more humid conditions beginning about 23 ka BP (reflected by lithologic data), followed by a transition toward wetter conditions and higher lake levels between 15 and 14 ka BP (indicated by our elemental data).

The elemental record and magnetic susceptibility signal in NP04-KH3 shows a correlation with some parts of the global Marine Oxygen Isotope chronology for Marine Isotope Stage (MIS) 3 through 1 (Fig. 12). Increased elemental concentrations and magnetic susceptibility from 60 to 32 ka BP correspond roughly in timing to MIS 3, followed by a transition to cooler conditions during MIS 2. However the transgression of Lake Tanganyika at the Kalya horst occurs around 18 ka BP and precedes the global transition from MIS 2 to MIS 1 at 13 ka. This record may support finding of early southern tropical climate change during the last glacial termination from other continents, such as evidence of early southern tropical deglaciation in the Andes (Seltzer et al., 2002).

Our record suggests key climate transitions at 50, 32, 23, and ~15 ka BP. Most aspects of this history correlate well with rainfall records derived from pollen assemblages in the Burundian highlands and southern Lake Tanganyika (Mpulungu) (Fig. 12; Bonnefille and Chalie, 2000) and other data from the Tanganyika basin. However, insolation at 4°S does not correspond strongly with our reconstructed environmental history from the Kalya horst indicating that the forces influencing the climate changes in the region are

more complex than solely astronomical forcings (e.g. Barker and Gasse, 2003). Certain aspects of the latest Pleistocene water balance history of Lake Tanganyika show correlations to high-latitude climate phenomena: our inference for peak aridity at ~23 ka BP is roughly in phase with maximum high-latitude ice sheet extent, and the arid–humid transition at ~15 ka BP, widespread, though perhaps not universal, in tropical Africa, is well-correlated with abrupt climate changes in Greenland (Stager et al., 2002). However, other aspects of our inferred climate history, such as the striking transition toward more arid conditions at ~32 ka BP, does not appear clearly correlated with global ice volume or high-latitude temperature, and may instead reflect changes in tropical sea surface temperatures and circulation (e.g. Barker and Gasse, 2003). New, long records of tropical African rainfall and local sea surface temperatures are needed to further examine the history of climate changes prior to the LGM.

5. Conclusions

Geochemical and sedimentological records from a core collected in central Lake Tanganyika provide a new record of climatic variability for East Africa over the last 60 ka BP. Geochemical and sedimentological analyses have proven useful at Lake Tanganyika for examining dramatic changes in a lakes' chemistry resulting from both weathering variability related to climate and variability of upstream watershed inputs from the volcanic/hydrothermal sources of Lake Kivu.

We have found evidence for relatively intense weathering and humid conditions (comparable to modern P/E) for the interval 60–50 ka BP. After 50 ka BP and until 32 ka BP weathering records suggest that climate conditions became somewhat drier, although still moderately humid. This was followed by a period of aridity between 32 and 14 ka BP, during the same period of the Last Glacial Maximum. Our findings suggest a rapid transition to the highly arid conditions of the LGM at 32 ka BP, earlier than some previously published estimates for the region, and a phased transition out of the arid conditions of the LGM.

The overflow of Lake Kivu into Lake Tanganyika is recorded in the sedimentary record in the Kalya region by increases in elemental concentrations and decrease in $^{87}Sr/^{86}Sr$ ratios. The addition of Lake Kivu saline/hydrothermal waters starting ~10 ka BP completely altered metal concentrations in Lake Tanganyika, affecting the geochemical signal of humid vs. arid intervals. The overflow of saline/metal rich water from Lake Kivu into Tanganyika appears to have been punctuated by an arid period from 8–6.8 ka BP, when Lake Kivu would have been a hydrologically closed basin. We propose that there were at least two distinct periods of overflow from Lake Kivu indicated by increased elemental concentrations and decreased $^{87}Sr/^{86}Sr$.

The paleoclimate record of the Lake Tanganyika basin recorded at the Kalya horst site over the past 60 ka BP shows striking similarities to the marine isotope record of glacial ice volume, and much less correspondence to precessionally driven insolation forcing over the same time interval. The onset of maximal aridity associated with the LGM appears to have been abrupt in the Kalya region, whereas the termination of aridity was more gradual.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.palaeo.2007.04.003.

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